

## Neoproterozoic chemostratigraphy and correlation of the Port Askaig glaciation, Dalradian Supergroup of Scotland

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**Abstract:** The  $\delta^{13}\text{C}$  of preglacial carbonates beneath the glaciogenic Port Askaig Formation falls from +5‰ to negative values, while least-altered  $^{87}\text{Sr}/^{86}\text{Sr}$  values are close to 0.7067. Postglacial ‘cap carbonates’ of the Bonahaven Formation begin with negative  $\delta^{13}\text{C}_{\text{carb}}$  values, jumping sharply to +11.7‰ near the top, which we correlate with excursions above the Elbobreen–Wilsonbreen glaciations of Svalbard and putative Sturtian (not Varanger) glaciations elsewhere, dated at *c.* 720 Ma. Much of the succeeding Terminal Neoproterozoic is missing in NE Svalbard and East Greenland, raising doubts about their use for global Sr and C isotope curves in the Neoproterozoic.

**Keywords:** Neoproterozoic, Dalradian, glaciation, chemostratigraphy.

The later Neoproterozoic arguably contains some of the most extreme climatic fluctuations in Earth history, from snowball earth to greenhouse conditions, with important implications for biosphere evolution (e.g. Hoffman *et al.* 1998). At present, however, there is no agreement on the number or extent of these glaciations: some recognize four, none of which were global (Kaufman *et al.* 1997) while others accept only two global glaciations (Kennedy *et al.* 1998). Distinction between these scenarios is hindered by the paucity of reliable chronometric constraints and biostratigraphic data (e.g. Brasier *et al.* 2000). Associated carbonates reveal large swings in  $\delta^{13}\text{C}$  values from highly positive values in the interglacials, falling to negative values during and/or following the glacial intervals, and these can be used to characterize particular time intervals (e.g. Knoll *et al.* 1986; Brasier *et al.* 1996; Kaufman *et al.* 1997; Kennedy *et al.* 1998), although they cannot provide unique signatures.

A major concern for Neoproterozoic stratigraphy, therefore, is completeness of the geological record: where and how long are the large time gaps? Without this information, it is difficult to establish a reliable chemostratigraphic or biostratigraphic scale. Strontium isotopes in carbonates may help here, if they are well-preserved, because there is evidence for a long-term trend towards more radiogenic  $^{87}\text{Sr}/^{86}\text{Sr}$  values between *c.* 800 and 500 Ma (e.g. Derry *et al.* 1989, 1992; Brasier *et al.* 1996; Shields 1999).

The Dalradian Supergroup of Scotland (Fig. 1) is one of the thickest (<25 km) and potentially one of the most complete Neoproterozoic–Cambrian successions to be found anywhere. It contains an important U–Pb zircon age of  $595 \pm 4$  Ma from a submarine keratophyre in the Tayvallich subgroup (Halliday *et al.* 1989), which lies up to 12 km above the Port Askaig Tillite (Harris *et al.* 1994). It has also been compared with the Laurentian successions of NE Svalbard and East Greenland, which have played a conspicuous role in the development of reference curves for carbon and strontium isotope stratigraphy in the Terminal Neoproterozoic (Knoll *et al.* 1986; Derry *et al.* 1989, 1992; Kaufman *et al.* 1997; Jacobsen & Kaufman 1999). The Dalradian succession differs from these in having a greatly expanded thickness of post-glacial Neoproterozoic sediments plus volcanics (Fig. 1; Soper 1994; Fairchild & Hambrey 1995).

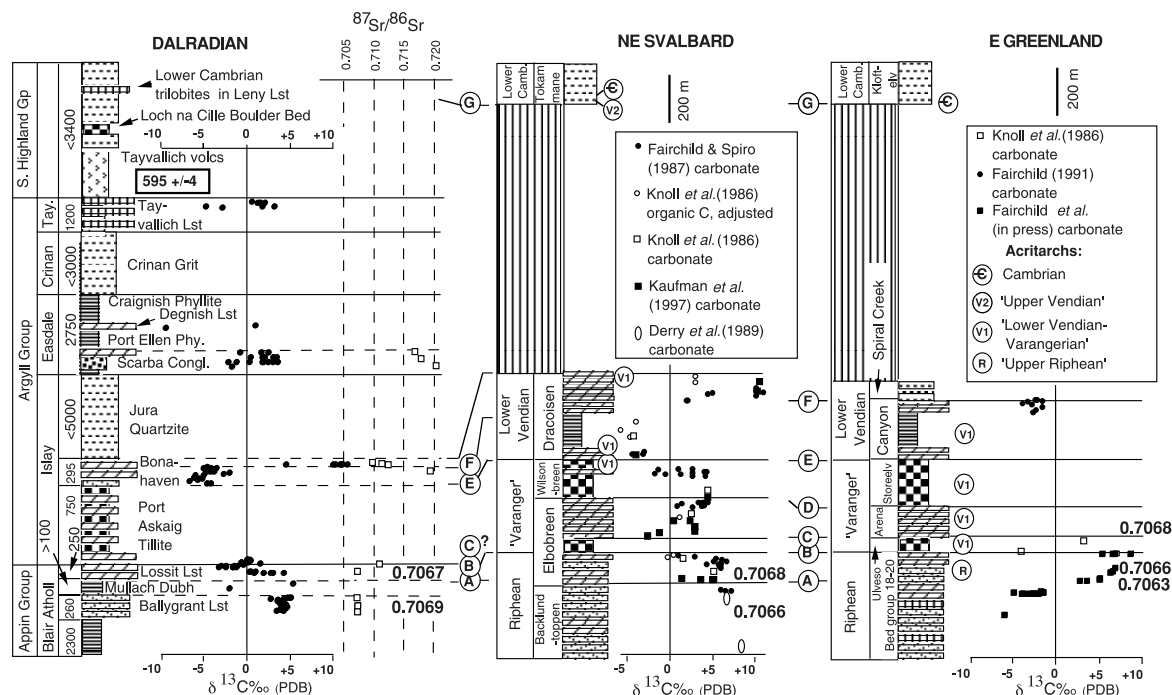
In this paper, we outline the carbon and strontium isotopic character of carbonates in the Dalradian Supergroup of Scotland with the aim of testing their potential for correlation of the Port Askaig glaciation with those in NE Svalbard and elsewhere.

### Geological setting

Sediments and metasediments of the Dalradian Supergroup are traceable along strike from NW Ireland to NE Scotland (Anderton 1985; Harris *et al.* 1994). Some 7 km of arkosic clastics of the Grampian Group, at the base of the supergroup, have been interpreted as either synrift or foreland basin deposits, laid down before *c.* 806 Ma (e.g. Soper 1994; Prave 1999a). The succeeding Appin Group (Fig. 1; *c.* 4 km thick) consists largely of sedimentary cycles in which argillites (now mainly phyllite or schist) pass up into shallower water quartzite or dark platformal limestones.

The Port Askaig Tillite (<750 m) at the base of the Argyll Group, consists largely of glaciomarine facies (Eyles & Eyles 1983). Abundant dolomitic clasts within diamictites near the base suggest local erosion of underlying dolostones, while nordmarkite clasts in feldspathic sandstone diamictites near the top indicate erosion of crystalline basement to the south (Anderton 1980). Deglaciation was followed by deposition of a thick ‘cap carbonate’ (the Bonahaven Formation, <295 m) which, along with associated quartz arenites, was laid down in peritidal to shallow subtidal conditions (Fairchild 1980). The succeeding Jura Quartzite, which may be up to 5 km thick, was deposited under tidal and storm-dominated shelf conditions in fault-bounded basins, during which increasing subsidence and depositional rates were broadly kept in balance (Anderton 1985).

Black shales of the Jura Slate and Easdale Slate, and turbidites plus mass flow deposits of the Scarba/Benderloch Conglomerate, together may indicate a phase of rapid basin deepening in response to major fault block movements (Anderton 1985). The upward passage from the Easdale Slate to the Degnish Limestone is gradational, and may indicate a progressive shallowing. Filling of this basin during deposition



**Fig. 1.** Summary of Dalradian stratigraphy showing the carbon (black dots) and strontium (white squares) isotope data from Islay, Jura and Argyll obtained in this paper, compared with that from NE Svalbard and East Greenland (after Knoll *et al.* 1986; Derry *et al.* 1989, 1992; Kaufman *et al.* 1997; Fairchild 1991, Fairchild *et al.* in press). A–F, event stratigraphic tie lines between NE Svalbard and East Greenland (after Fairchild & Hambrey 1995) and Scotland (suggested herein): A, deepening event in pre-glacial carbonate platform; B, erosional to conformable sequence boundary at base of lower glacial deposits; C, rapid deglaciation and transgression; D, erosional to conformable sequence boundary at base of upper glacial deposits; E, rapid deglaciation and transgression; F, highstand, regressive facies; G, Lower Cambrian transgression. Acritarch assemblages from Svalbard and Greenland after sources in Knoll & Swett (1987).

of the Craignish Phyllites is indicated by evidence for evaporites (Anderton 1985). A second major cycle of fault-bound subsidence produced submarine fans of the Crinan Grit and turbiditic carbonates of the Tayvallich Limestone. Continued subsidence led to deposition of thick turbiditic sandstones of the Southern Highland Group, including localized mafic extrusives such as the Tayvallich Volcanics (Harris *et al.* 1994). The Leny Limestone of Perthshire, which comprises thin bedded turbiditic carbonates, arguably belongs to the uppermost part of this group (Tanner 1995).

### Age of the Dalradian

Correlation of the Dalradian is hampered by the rarity of biostratigraphic markers. The trilobite *Pagetides* sp. provides a late early Cambrian age for the Leny Limestone at the top of the succession (Pringle 1940). A post-Varanger age for the Argyll Group has been inferred on the basis of putative metazoan trace fossils. Re-examination by us of *Neonereites* markings reported by Brasier & McIlroy (1998) shows, however, that these could arguably have been produced by sedimentary processes, while carbonate clasts at Benderloch reveal that the *Cloudina*-like forms of Conway Morris (1998) may be parts of composite coated grains or oncolites. The Precambrian–Cambrian boundary (*c.* 543 Ma; Grotzinger *et al.* 1995) must, however, lie somewhere beneath the Leny Limestone and above lavas dated at  $595 \pm 4$  Ma (Halliday *et al.* 1989).

Until recently it has been assumed, on the basis of lithostratigraphy, that the Port Askaig Tillite correlates with the twin Varanger tillites of Norway, the Elbobreen and

Wilsonbreen tillites of NE Svalbard, plus the Ulveso and Storeelv tillites of East Greenland (e.g. Soper 1994; Harris *et al.* 1994). These, in turn, have been correlated by lithostratigraphy with the Marinoan glaciation of south and central Australia, and by carbon isotope stratigraphy with the Icebrook glaciation of NW Canada, estimated at *c.* 590–564 Ma old (e.g. Kaufman *et al.* 1997; Saylor *et al.* 1998). Kennedy *et al.* (1998) have argued, however, that the upper 'Varanger' glacial in Norway, Svalbard and Greenland is Marinoan, whereas the lower 'Varanger' glacial is Sturtian (*c.* 723 Ma, see Brasier *et al.* 2000). Prave (1999a) has gone further and suggested a wholly Sturtian age for the Port Askaig glaciation, since it lies up to 12 km below strata dated at 595 Ma (Fig. 1). This would mean that Marinoan/younger Varanger glaciations could lie higher in the Dalradian, at the level of the Loch na Cille Boulder Bed (Prave 1999a) and/or glacially-influenced sediments in the Macduff Slates (see Stoker *et al.* 1999), both in the Southern Highland Group (Fig. 1).

### Neoproterozoic–Cambrian isotope stratigraphy

A decade of intensive study into carbon and strontium isotopes from carbonate rocks of the Neoproterozoic–Cambrian means that chemostratigraphy can provide a way to test the relative age of the Port Askaig Tillite. Most important are reference sections in which the stratigraphic succession is unambiguous, the major sedimentary breaks are acknowledged, the palaeontological datum points (including the Precambrian–Cambrian boundary) are clear, and the carbonates contain well-preserved  $\delta^{13}\text{C}$  and (most importantly)  $^{87}\text{Sr}/^{86}\text{Sr}$  signatures. In our view,

the successions of SW Mongolia and Siberia (Brasier *et al.* 1996), the Mackenzie Mountains of Canada (Kaufman *et al.* 1997), the Kalahari craton of Namibia (Grotzinger *et al.* 1995) and Oman (Brasier *et al.* 2000) must play key roles. Unfortunately, the ages of Neoproterozoic successions from NE Svalbard, East Greenland, and the Otavi craton of Namibia remain controversial (see Kaufman *et al.* 1997; Kennedy *et al.* 1998; Jacobsen & Kaufman 1999).

We infer that the older glaciations (including the Rapitan of Canada, Sturtian of Australia and Maikhan Uul of Mongolia) are bracketed by  $^{87}\text{Sr}/^{86}\text{Sr}$  signatures in carbonates of *c.* 0.7067, and followed by extremely positive  $\delta^{13}\text{C}$  values of  $>+10\text{‰}$  that are otherwise unmatched in the last 1000 Ma. The younger glaciations (including the Icebrook of Canada, Marinoan of Australia and Fiq of Oman) are followed by  $^{87}\text{Sr}/^{86}\text{Sr}$  values that rise rapidly from *c.* 0.7072 to 0.7084 and by  $\delta^{13}\text{C}$  values that do not rise above  $+8\text{‰}$ . The latter interval is also characterized by the Pertatataka acritarch microflora, and this in turn is followed by unequivocal Ediacara and *Cloudina* biota (e.g. Brasier *et al.* 2000).

120 samples were collected for chemostratigraphy from carbonate outcrops in the Dalradian of Islay, Jura and the nearby Argyll mainland, where the metamorphic grade is relatively low and the microfabrics are well-preserved (e.g. Fairchild 1985). All samples were thin-sectioned, stained with potassium ferricyanide and alizarin red-S, and studied microscopically for evidence of major diagenetic (e.g. ferroan) carbonate phases. Where available, fine-grained carbonate (mainly comprising micritic or microsparitic mudstone) was selectively drilled from the most homogeneous regions of the rock slice, to form a powder for oxygen and carbon isotopic analysis using the techniques described in Brasier *et al.* (1996). 22 samples were selected on the basis of sample homogeneity and stable isotope values, for strontium isotope analysis in Strasbourg. Sample chips were dissolved during 24 hours in 100% glacial acetic acid, with distilled water added dropwise until reaction was evident. After standard Sr concentration procedures, analysis was carried out on a VG Sector multicollector in dynamic mode; the standard NBS 987 averaged 0.710286 ( $n=3$ ) over the period of measurement (cf. 0.710201, Fairchild *et al.* in press; 0.710241, Derry *et al.* 1989, 1992; and 710254, Brasier *et al.* 1996). Data discussed herein are normalized to an NBS 987 value of 0.71025. The same leachates were evaporated and analysed as 1 M  $\text{HNO}_3$  solutions for their trace element chemistries using ICP-AES (Ca, Mg, Sr, Ba, Mn, Fe, K, Na) and ICP-MS (Rb, Sr). Rb concentrations were too low for  $^{87}\text{Rb}$  decay to alter significantly the limestone  $^{87}\text{Sr}/^{86}\text{Sr}$  data.

## Results

A summary of our chemostratigraphic data from the Dalradian (Fig. 1, Table 1) indicates a fall in maximum  $\delta^{13}\text{C}$  values from *c.*  $+5\text{‰}$  in the Ballygrant Limestone and Mullach Dubh Phyllite, towards negative values in the Persabus dolostones of the Lossit Formation, beneath the Port Askaig Tillite. Although this trend appears to have been accentuated by diagenesis, particularly in ferroan dolostones of the Mull of Oa section where  $\delta^{18}\text{O}$  values fall below  $-9\text{‰}$ , it is also seen in shallow water non-ferroan dolostones from Loch Lossit and Keills in which  $\delta^{18}\text{O}$  values are  $>-5\text{‰}$ . This suggests that the negative trend beneath the Port Askaig glaciation could reflect a global change in seawater chemistry. Dolostone interbeds and diamictites within the lower part of the Port Askaig Tillite have carbon isotopic values close to those from the underlying Lossit dolostones, consistent with their derivation from these beds by glacial erosion.

Members 1 and 3 of the Bonahaven Formation 'cap carbonate' (Fig. 1) show a fall in  $\delta^{13}\text{C}$  of more than  $2\text{‰}$ . Oxygen

isotope values are comparable with those from the underlying Port Askaig Tillite and Lossit Formation. Although ferroan dolomicrite fabrics and strong C–O covariance ( $r^2=0.7$  to  $0.9$ ) may be taken to suggest that the more negative carbon isotopic signals have been reset along a mixing line, negative values are also found within less-altered non-ferroan dolomite and calcitic limestone components. The dramatic rise to  $\delta^{13}\text{C}$  values around  $+11.7\text{‰}$  in Member 4 (Fig. 1) is unaccompanied by any major shift in oxygen isotopes, but does mark a switch from Fe- and Mn-rich to Fe- and Mn-poor dolomites. This is one of the most extreme carbon isotopic swings in the geological record and therefore provides a potential marker for correlation. A single sample at the top of Member 4 has lower values ( $+4.5\text{‰}$ ).

Carbon isotopic values from carbonates of the Argyll Group (i.e., the Scarba Conglomerate, Port Ellen Phyllite, Degnish Limestone and Tayvallich Limestone) reveal a relatively large spread (Fig. 1, Table 1) which may, in part, reflect the influence of allochthonous carbonate sources. The  $\delta^{13}\text{C}$  maxima do not much exceed  $+3\text{‰}$ , however, and this may be close to seawater values for the interval.

Strontium isotopic ratios of  $>0.7096$  ( $n=10$ ) in the Argyll Group limestones and dolostones are accompanied by positive correlations between  $^{87}\text{Sr}/^{86}\text{Sr}$  and Mn/Sr (Table 2), indicating diagenetic alteration of Sr isotope signals during diagenesis (cf. Brasier *et al.* 1996). Appin Group limestones are better preserved, having high Sr concentrations ( $>1000$  ppm) and low Mn/Sr ratios ( $<0.2$ ), yielding values of between 0.70691 and 0.70725 ( $n=6$ ) for the Ballygrant Limestone, falling to between 0.70669 and 0.70720 for the Persabus Member of the Lossit formation, a few tens of metres beneath the Port Askaig Tillite (Fig. 1; Table 2). The lowest  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios correspond to samples with the lowest Mn/Sr ratios ( $<0.1$ ), implying that these values provide maximum constraints on seawater  $^{87}\text{Sr}/^{86}\text{Sr}$ .

## Discussion

Both the carbon and strontium isotope results are consistent with previously suggested comparisons between the Appin Group to Bonahaven Formation of the Scottish Dalradian, the Backlundtoppen to Dracoisen formations of NE Svalbard, and bed groups 18–20 up to the Spiral Creek Formation of East Greenland (Fig. 1; e.g., Fairchild 1989; Fairchild & Hambrey 1995; Soper 1994). In each area, carbonates beneath the lower glacial unit yield low, least-altered  $^{87}\text{Sr}/^{86}\text{Sr}$  values close to 0.7067 (Fig. 1), while carbonates above the upper glacial unit, in the Bonahaven and Dracoisen formations contain a distinctive carbon isotopic marker with values of  $+11\text{‰}$ . Above this in Scotland, there is evidence for increasing platform instability and rapid subsidence, followed by marked deepening and volcanic activity. Conversely, in NE Svalbard and East Greenland, both acritarchs and sedimentology suggest several prolonged breaks (e.g. Knoll & Swett 1987; Fairchild 1991) implying that this part of the Laurentian margin experienced non-deposition or erosion.

Carbon isotopic values of  $>+10\text{‰}$  provide a distinctive marker for the interval between the Sturtian and Marinoan/Varanger glaciations (Brasier *et al.* 1996; Jacobsen & Kaufman 1999). Hence  $\delta^{13}\text{C}$  values as high as  $+11.7\text{‰}$  in the Bonahaven Formation, when taken together with  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios as low as 0.7067 beneath the Port Askaig Tillite, suggest correlation with the older, Sturtian/Rapitan glaciations rather than with

**Table 1.** Carbon and oxygen isotope data from Dalradian carbonates

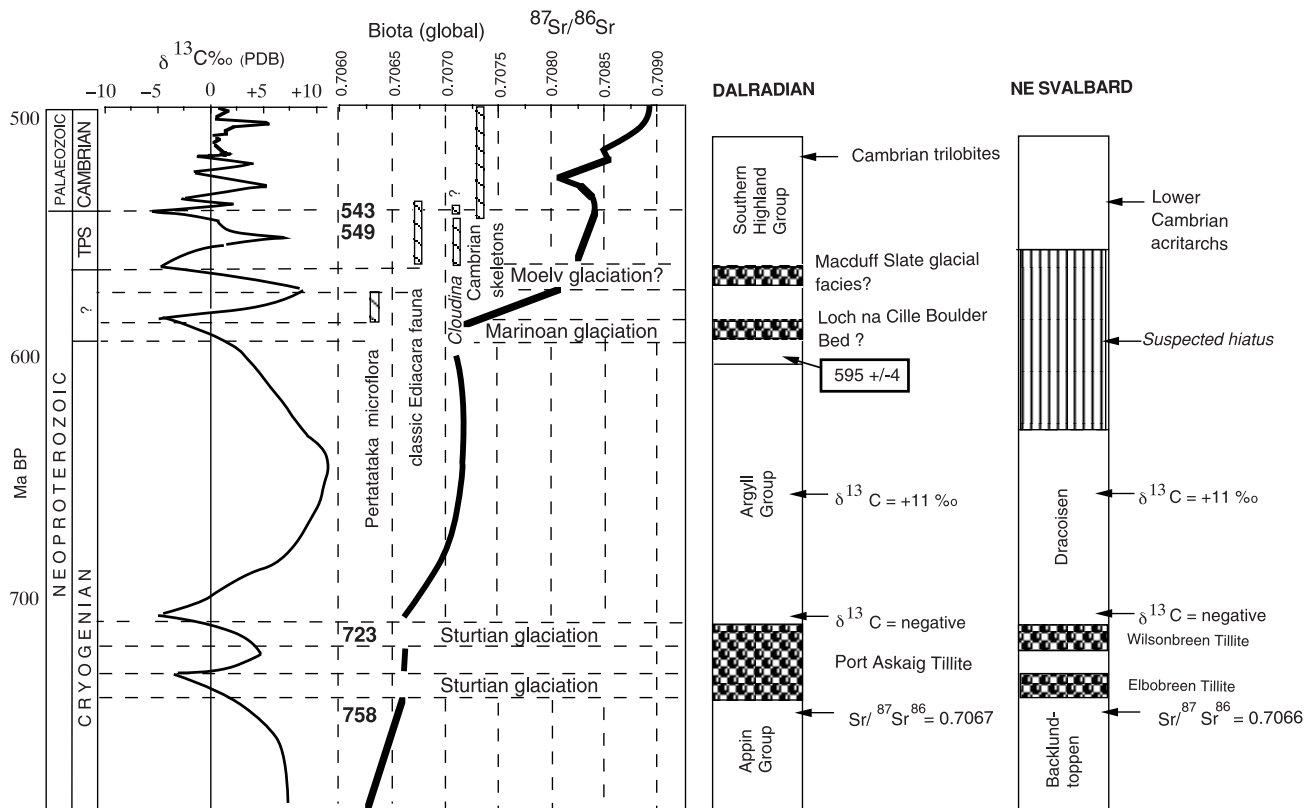
Sample	Height	Lithology	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	Sample	Height	Lithology	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$
Tayvallich Lst:					Pt Askaig Fm:				
Ben 8	4742	fds qtz py	0.5	-4.9	LL1	500	fdms qtz	-0.7	-10.3
Ben 7	4738	fds qtz py	1.5	-4.5	LL3	500	dm dms qtz	-0.1	-8.9
Ben 6	4733	fds qtz	2.0	-1.2	LL2	499	dm dms qtz	0.3	-8.2
Ben 5	4732	fds qtz py	1.3	-1.9	LL4	497	dms	0.1	-8.8
Ben 4	4692	fds qtz py	3.1	-4.6	LL5	495	qtz dms mx	0.2	-7.5
Ben 3	4691	cs ds qtz py	1.8	-7.3	Lossit Fm:				
Ben 2	4690	fds qtz py	-4.8	-4.6	LL6	492	fdms qtz	0.4	-6.8
Ben 1	4687	fds qtz py	-2.8	-5.2	LL7	482	dm qtz	1.5	-4.5
Degnish Lst:					LL9	472	fcs qtz	-0.4	-0.8
Deg 2	3666	fcs qtz	0.9	-15.8	LL8	462	dms ds	-1.7	-6.2
Deg 1	3665	qtz fcs	-9.5	-14.5	OA2	459	fdms qtz mic	-1.1	-11.1
Scarba Congl.:					OA3	458	fdms qtz	-1.4	-11.2
Kin 1	2665	qtz phy dm	1.7	-12.3	OA3a	457	fdms qtz	-3.0	-12.3
Kin 2	2665	qtz phy fdm py	-0.8	-14.4	OA4	456	fdms qtz	-2.4	-12.1
Kin 3	2664	fdm qtz	2.5	-14.8	OA5	455	fds fdms	-1.7	-12.0
Bend 3a1	2614	fcs ic qtz	3.5	-14.2	OA6	454	fdms ds	-1.6	-10.7
Bend 3a2	2614	fcs ic qtz	3.0	-14.3	OA7	453	dms qtz	-1.3	-10.0
Bend 3b	2614	qtz fcs ic	1.8	-13.8	OA8	452	fdm qtz mic	-2.5	-9.3
Bend 3c1	2614	qtz fcs	0.4	-13.6	OA9	451	fdm fdms mic	-3.4	-9.5
Bend 3c2	2614	qtz fcs onc	2.0	-13.9	Pe 9	399	dm fds	0.2	-5.2
Bend 2a	2564	qtz phy fcs	-2.2	-9.5	Pe 8	397	dm fds	0.4	-4.7
Bend 2b	2564	fcs onc	3.5	-13.3	Pe 7	395	dm ds	0.9	-5.5
Bend 2c	2564	qtz fcs	1.9	-12.2	Pe 6	393	dm fdm ds	0.8	-6.6
Bend 2d	2564	cs fcs	0.3	-12.1	Pe 5	391	cs qtz	3.0	-4.2
Bend 2e	2564	qtz phy fcs	3.0	-15.0	Pe 4	389	cs qtz	2.8	-5.5
Bend 2f	2564	qtz fcm	-1.2	-11.8	Pe 3	387	cms qtz	1.9	-10.2
Bend 2 g	2564	fcs onc cs	2.4	-14.1	Pe 1	383	cms qtz	4.3	-6.6
Bend 2 h	2564	cs onc fcs	1.7	-14.4	Mullach Dubh:				
Bend 1	2514	fcs qtz	-1.8	-10.3	OA10	234	fems cs	5.3	-12.3
Bonahaven Fm:					OA11	233	dms qtz	-2.0	-9.4
Mem 4 Bd 16	1513	cms fcm qtz	4.5	-5.9	Ballygrant Lst:				
Bd 17	1512	dms qtz	11.0	-9.9	Ba 28	133	cms cs ic	5.0	-7.4
Bd 18	1511	dms ds py	10.8	-9.1	Ba 27	132	cms cs ic mic	4.9	-7.1
Bd 19	1510	dms cs qtz py	11.7	-9.7	Ba 26	128	cms cs ic	4.6	-7.2
Bd 13	1508	dms qtz py	10.3	-10.5	Ba 25	123	cms cs ic	4.3	-7.3
Bd 14	1506	dm cs qtz py	11.0	-10.1	Ba 24	118	cms ds	2.7	-10.3
Bd 15	1505	dm qtz py	10.0	-10.8	Ba 23	113	cms ds	3.6	-6.3
Mem 3 Bd 12	1480	fdm qtz gy?	-4.7	-11.5	Ba 22	108	dms ds	3.5	-4.2
Bd 11	1475	fdm qtz	-5.0	-11.7	Ba 21	103	cm cms ds	3.6	-7.1
Bd 10	1470	dm cs qtz py	-4.5	-11.6	Ba 20	98	cm dm ic	4.0	-6.9
Bd 9	1469	dms cs qtz	-4.2	-10.5	Ba 19	93	cm dm ic	3.7	-7.6
Bd 8	1468	fdm cs qtz	-3.5	-7.4	Ba 18	88	cms dm ic oo	4.3	-9.5
Bd 7	1456	fdm qtz py	-3.8	-8.9	Ba 17	83	cm dm ic	3.9	-6.1
Bd 6	1455	dm fdms qtz py	-4.5	-11.0	Ba 16	78	cm dm ic	4.1	-7.7
Bd 5	1450	dm fdm qtz	-4.7	-10.4	Ba 15	73	cm cms	3.9	-9.5
Bd 4	1447	dm fdm qtz	-4.6	-9.9	Ba 1	63	cm dm ic	4.5	-5.6
Bd 3	1442	qtz fdm py	-4.7	-10.3	Ba 2	58	cm cms	4.2	-8.0
Bd 2	1439	dm qtz py	-3.8	-9.0	Ba 3	53	cms dm ic	4.4	-7.3
Bd 1	1437	fdm qtz py	-4.2	-9.9	Ba 4	48	cm dm ic	4.3	-7.6
Bun 1	1419	dm fdm qtz py	-2.0	-7.3	Ba 5	43	cm dm ic	4.0	-6.1
Bun 2	1409	dm fdm qtz py	-2.8	-8.1	Ba 6	38	cm dm ic	4.6	-6.0
Bun 3	1397	dm fdm qtz py	-5.1	-10.6	Ba 7	34	cm dm ic	4.4	-6.0
Bun 4	1393	dm cs py	-6.1	-10.9	Ba 8	29	cm dm ic	4.5	-6.8
Bun 5	1378	dm fdm qtz py	-6.2	-10.4	Ba 9	24	cms oo	4.4	-8.3
Bun 6	1364	dm fdm qtz py	-5.2	-9.4	Ba 10	19	cms oo	4.4	-6.2
Bun 7	1362	fdm oo fdm qtz	-5.6	-10.8	Ba 11	15	cm dm ic	4.0	-6.3
Bun 8	1359	fdms dms cs	-5.9	-11.2	Ba 12	10	cm dm ic	3.9	-7.3
Bun 11	1354	fdm qtz py	-5.8	-11.5	Ba 13	5	cm cms	3.4	-10.0
Bun 12	1344	dm fdm qtz	-6.7	-11.2	Ba 14	0	cms dm ic	4.3	-6.8
Mem 1 Bun 10	1314	dms qtz	-4.6	-12.0					
Bun 9	1312	dms sph	-4.3	-10.5					

All heights (in metres) are referenced to zero at the base of the Ballygrant Limestone. Localities: Ballygrant Lst (Ba 1-28, NR 395658); Mullach Dubh Phyllite (OA 10, 280418; OA 11, NR 281421); Lossit Lst (Pe 1-9, NR 422687; LL6-9, NR 414647; OA 1-9, NR 282413); Port Askaig Tillite (LL1-5, NR 414647); Bonahaven Fm (Bun 1-11, NR 423733; Bd 1-19, 407793); clasts in Scarba/Benderloch Conglomerate (Bend 1, NM 902380; Bend 2, NM 904385; Bend 3, NM 906386); top of Scarba Congl./base of Port Ellen Phyllite (Kin 1-3, NR 706992); Degnish Lst (Deg 1, NM 777123, Deg 2 float, NM 783123); Tayvallich Lst (Ben 1-8, NR 867713). Mem, member. Lithologies: c, calcitic; d, dolomitic; f, ferroan; gy, gypsum pseudomorphs; ic, carbonate intraclasts; m, micrite; mic, mica flakes; ms, microspar; onc, oncolitic; oo, oolitic; phy, phyllitic; py, pyritic; qtz, quartz sand/silt; s, spar; sph, phengite spherules.



**Table 2.** Strontium isotopes and analytical data obtained from selected Dalradian carbonates (see Table 1)

Sample	Insol. residue	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	Sr/Ca	K/Ca	Mg/Ca	Mn/Sr	Fe/Sr
Bend 3a2	33.6%	<0.00010	0.716600	0.00402	0.0006	0.01030	1.39	4.67
Bend 2 g	4.5%	<0.00010	0.717561	0.00289	0.0035	0.00888	1.02	5.16
Bend 2H	4.2%	<0.00010	0.720105	0.00694	0.0003	0.03460	1.25	3.76
Bd 19	2.0%	0.00109	0.709573	0.00101	0.0028	0.318	0.45	2.77
Bd 18	22.0%	0.00239	0.709606	0.00101	0.0031	0.309	0.35	2.44
Bd 17	3.0%	0.00294	0.709768	0.00123	0.0029	0.284	0.44	2.74
Bd 15	7.0%	0.00308	0.712116	0.00092	0.0035	0.318	0.92	4.54
Bd 14	12.8%	0.00056	0.711049	0.00075	0.0022	0.268	0.67	4.33
Bd 12	20.4%	0.00192	n.a.	0.00121	0.0027	0.314	5.20	98.60
Bun 6	18.7%	0.00132	0.719194	0.00136	0.0034	0.347	4.25	74.70
OA3	5.8%	<0.0006	n.a.	0.00060	0.0025	0.305	2.69	34.75
OA6	4.0%	<0.0009	0.710789	0.00075	0.0032	0.372	1.88	23.84
OA7	5.2%	0.00023	n.a.	0.00160	0.0031	0.342	1.13	8.97
Pe 9	5.0%	0.00316	n.a.	0.00130	0.0036	0.305	4.23	12.40
Pe 7	1.0%	<0.00010	0.706693	0.00621	0.0002	0.00518	0.068	0.56
Pe 3	5.7%	<0.00010	0.707055	0.00421	0.0001	0.00656	0.042	0.32
Ba 28	9.8%	<0.00010	0.707200	0.00467	<0.0001	0.00692	0.034	0.66
Ba 27	4.0%	<0.00010	0.706911	0.00513	0.0003	0.00534	0.032	0.49
Ba 1	8.1%	<0.00010	0.707252	0.00294	0.0003	0.00634	0.169	0.49
Ba 2	2.4%	0.00047	0.707006	0.00473	0.0012	0.00659	0.036	0.32
Ba 13	10.5%	<0.00010	0.707035	0.00363	<0.0001	0.00501	0.189	0.41
Ba 14	1.0%	<0.00010	0.707219	0.00536	0.0005	0.00860	0.044	0.17



**Fig. 2.** Carbon and strontium isotopic curves for the Neoproterozoic–Cambrian, based largely on data from SW Mongolia, NW Canada and Oman (from sources cited in the text). Major U–Pb zircon dates shown in bold (from sources in Brasier *et al.* 2000). At right is shown a suggested time span for the Dalradian and NE Svalbard successions.

the younger, Marinoan/Icebrook and Varanger glaciations (Fig. 2). An older, Sturtian, age for these glaciations is also consistent with a U–Pb zircon date of 595 Ma from the Tayvallich volcanics, and with the inferred presence of glacial

deposits higher in the Southern Highland Group (Stoker *et al.* 1999; Prave 1999a).

We therefore question the commonly held assumption of a younger Varanger/Marinoan age for the twin glaciations of

NE Svalbard and East Greenland (e.g. Jacobsen & Kaufman 1999) and agree with Prave (1999a) that these glaciations have Sturtian affinities. If so, then we must also question whether post-Sturtian and post-Marinoan 'cap carbonates' invariably had broadly time-specific lithofacies (i.e., bituminous limestones after Sturtian; dolostones after Marinoan; cf. Kennedy *et al.* 1998) and caution against the use of 'cap' lithofacies for global chronostratigraphy (cf. Prave 1999b). It seems more likely to us that 'cap' lithofacies and their carbon isotopic profiles were influenced by water depth and regional palaeoceanography.

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